

1                   Annual and Interannual Variations of Phytoplankton Pigment Concentration  
2   and Upwelling Along the Pacific Equator  
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## ABSTRACT

The following variables along the Pacific equator from 145°E to 95°W were employed: surface-layer phytoplankton pigment concentrations derived from Nimbus-7 Coastal Zone Color Scanner (CZCS) measurements of ocean color radiances; vertical velocities simulated at the 90-m bottom of the euphotic layer from a wind-driven ocean general circulation model; nitrate estimated from model simulated temperature. The upward flux of nitrate into the euphotic layer was calculated from the simulated vertical motion and nitrate. The CZCS-derived phytoplankton pigment concentration was uniform from 175°W to 95°W. Longitudinal profiles of upwelling, phytoplankton biomass, and 90-m nitrate flux were of different shape. The small annual cycles of the phytoplankton pigment concentration and nitrate flux were in phase: increased phytoplankton biomass was associated with increased upward nitrate flux, but the phase was not consistent with the annual cycles of the easterly wind or of the upwelling intensity. El Niño variation of phytoplankton pigment concentration was greater than the annual cycle. The substantially reduced phytoplankton pigment concentration observed during El Niño was associated with smaller upward nitrate flux. Phytoplankton biomass during non-El Niño conditions was not related to the nitrate flux into the euphotic layer.

## 1. INTRODUCTION

The Nimbus-7 Coastal Zone Color Scanner (CZCS) measurements of phytoplankton pigment concentration were approximately uniform along the Pacific equator from several hundred kilometers west of the Galapagos to the date line [*U. S. News and World Report* 1989]. The usual wisdom dictates that chlorophyll-*a* fluctuations mimicked sea surface temperature (SST) and mixed-layer depth variations. The isothermal and isopycnal mixed-layer depths are approximately 20 m near the Galapagos Islands at 90°W and about 150 m at the date line [*Colin et al.*, 1971]. Sea surface temperature is usually 5 °C lower near the Galapagos than at the date line [*Reynolds,*

1982]. The euphotic-layer depth, which was estimated from CZCS-derived phytoplankton pigment concentrations (see §3.1), was about 90 m throughout the length of the Pacific equator [Morel, 1988]. Application of the Sverdrup [1953] critical-depth mechanism suggests that chlorophyll-*a* would be higher where the euphotic-layer depth was greater than the mixed-layer depth, such as in the eastern equatorial Pacific, and *vice versa* in the western Pacific.

That the chlorophyll-*a* was higher, SST was lower, and mixed-layer depth was smaller on the equator than at several hundred kilometers north and south of the equator was known for many decades. Because the prevailing zonal direction of the surface wind along the equator is westward, upward motion or upwelling at the equator is thought to be a prominent physical mechanism responsible for the observed meridional distributions of SST and near-surface phytoplankton biomass. At the equator, zooplankton grazing of phytoplankton is considered to be of lesser importance than the physical dynamics [Barber, 1988]; however, at several hundred kilometers north and south of the equator, zooplankton grazing reduces the size of the phytoplankton population formed by equatorial upwelling [Vinogradov, 1981; Barber and Chavez, 1991].

A tenet of faith among oceanographers that upper-ocean vertical motions along the Pacific equator strongly influence the chlorophyll-*a* distribution will be examined herein. Vertical motion in the ocean for time scales longer than several days is very small [Stommel, 1964], and the difficult measurements are scarce, sporadic and intermittent [Halpern and Barber, 1987]. The vertical velocity will be described from ocean general circulation model (OGCM) simulations. This paper describes a search for a correlative relationship between the zonal distributions of phytoplankton biomass, upwelling, and nutrient content of the upwelled water along the Pacific equator.

## 2. METHODS

### 2.1. CZCS-Derived Phytoplankton Pigment Concentration

1       The National Aeronautics and Space Administration (NASA) Nimbus-7 satellite was  
 2       launched on 23 October 1978 into a sun-synchronous orbit having a nominal altitude of 955 km  
 3       and a repeat cycle of about 6 days. The CZCS sensor, which was one of eight instruments on the  
 4       spacecraft, continued operation until June 1986. The CZCS was a spatially imaging multi-spectral  
 5       scanner with an 825 m x 825 m footprint or pixel size and a cross-track swath of 1566 km. Each  
 6       swath overlaps the preceding one by about 25%.

7       The CZCS coverage was worldwide, but it was by no means uniform because the CZCS  
 8       was not intended to collect a global data set. The CZCS operated about 10% (sometimes up to  
 9       20%) of each orbit and data were recorded primarily in a broadcast mode via ground stations.  
 10      Geographical areas receiving the greatest coverage were the coastal and offshore regions of the  
 11      United States, western Mediterranean Sea, and Arabian Sea [McClain *et al.*, 1993]. In the tropical  
 12      Pacific, very large data gaps prevented the assembly of successive monthly composites of the  
 13      phytoplankton concentration for even one complete annual cycle.

14      The CZCS measured radiances emitted below the ocean surface and at the ocean surface in  
 15      six co-registered spectral bands: 1 (blue), 433 - 453 nm; 2 (green), 510 - 530 nm; 3 (yellow), 540  
 16      - 560 nm; 4 (red), 660 - 680 nm; 5 (near infrared), 700 - 800 nm; 6 (thermal infrared), 10500 -  
 17      12500 nm. Bands 1 - 3, which are in the visible region of the electromagnetic spectrum, are  
 18      corrected for atmospheric and surface effects to yield water-leaving radiances. Ratios of bands 1 -  
 19      3 water-leaving radiances are used to calculate phytoplankton pigment concentration. No universal  
 20      algorithm exists to convert CZCS radiances to phytoplankton pigment concentration, and the exact  
 21      model functions and magnitudes are issues of current investigations. The CZCS-derived  
 22      phytoplankton pigment concentration is determined to within a factor two in open-ocean regions  
 23      without comprehensive *in situ* calibration [Hickman *et al.*, 1991]. The uncertainty of CZCS  
 24      estimates of phytoplankton pigment concentration made coincidentally with *in situ* observations is  
 25      reduced to 30 - 40% [Hickman *et al.*, 1991].

26      This report uses phytoplankton pigment concentrations computed from CZCS measurements  
 27      processed at the NASA Goddard Space Flight Center [Feldman *et al.*, 1989]. The empirical bio-

optical algorithms related water-leaving radiances to CZCS-derived phytoplankton pigment concentration,  $C$ , are [Gordon *et al.*, 1983]:

$$C = 1.13 (Lw_{443}/Lw_{550})^{-1.71} \text{ for } C \leq 1.5 \text{ mg m}^{-3} \quad (1)$$

and

$$C = 3.33 (Lw_{520}/Lw_{550})^{-2.44} \text{ for } C > 1.5 \text{ mg m}^{-3}, \quad (2)$$

where  $Lw_{443}$ ,  $Lw_{520}$ , and  $Lw_{550}$  are the CZCS-measured radiances ( $\text{mW m}^{-2}$ ) in spectral bands 1, 2 and 3, respectively. The algorithm bifurcation at  $1.5 \text{ mg m}^{-3}$  was necessary because  $Lw_{443}$  becomes too small to accurately quantify with the 8-bit digitization interval of the CZCS sensor.

Balch *et al.* [1992] reported that  $C$  was about 20% lower than coincident *in situ* data for  $C < 1 \text{ mg m}^{-3}$ , which was the range of  $C$  for the equatorial Pacific (see §3.1). The minimum detectable value of the CZCS-derived phytoplankton pigment concentration is  $0.037 \text{ mg m}^{-3}$  per  $18\text{-km} \times 18\text{-km}$  pixel, which was the spatial resolution of the GSFC-archived CZCS image. The CZCS algorithm was virtually independent of water vapor variations because of the high atmospheric transmittances associated with the blue, green and yellow radiances. Wind speed variations did not influence the CZCS estimate of phytoplankton pigment concentration [Kanno, 1992] because  $C \approx 0.2 \text{ mg m}^{-3}$  (see §3.1) and the equatorial Pacific wind speed rarely exceeded  $10 \text{ m s}^{-1}$  [Halpern, 1988].

Aerosol content and intensity were approximately uniform along the Pacific equator from  $145^\circ\text{E}$  to  $95^\circ\text{W}$  because of the remoteness of the region from land.

The phytoplankton pigment concentration represents the sum of chlorophyll-*a* and phaeophytin-*a*, and is an index of phytoplankton biomass. The CZCS-derived values represent the average pigment concentration within the upper 20-30% of the euphotic layer for  $0.1 - 10.0 \text{ mg m}^{-3}$  pigment concentrations [Smith and Baker, 1978]. Within  $5^\circ$  of the Pacific equator from  $180^\circ$  to  $95^\circ\text{W}$ , the mean 1983 - 1990 chlorophyll concentration maximum ( $\approx 0.35 \text{ mg m}^{-3}$ ) was at 40- to 50-m depth [Barber and Chavez, 1991], which was approximately one-half the 90-m thickness of the euphotic layer (see §3.1). Thus, along the Pacific equator, the CZCS data did not penetrate the chlorophyll maximum.

## 2.2. OGCM-Derived Vertical Velocity

The four-dimensional OGCM was developed at the National Oceanic and Atmospheric Administration (NOAA) Geophysical Fluid Dynamics Laboratory (GFDL) by Bryan [1969] and Cox [1984], and modified for the tropical Pacific Ocean by Philander and Seigel [1985]. Numerous comparisons between the Philander/GFDL OGCM and *in situ* data [Philander, 1990] indicated that numerical simulation of complicated tropical oceanographic phenomenon was acceptable for studies of the physical insight into dynamics.

The Philander/GFDL OGCM, which was transferred to the CRAY Y-MP at the NASA Ames Research Center [Chao *et al.*, 1991], is based on the Reynolds equations of motions in spherical coordinates with the Boussinesq and hydrostatic approximations. Prognostic variables are temperature, salinity, and the zonal and meridional velocity components. The vertical velocity,  $w$ , is calculated from the continuity equation. The subgrid-scale turbulent fluxes are parameterized by a first-order turbulence closure scheme: the vertical eddy exchange coefficient is parameterized by an empirical formula [Pacanowski and Philander, 1981] of the Richardson number, and the horizontal eddy exchange coefficient is assumed to be a constant of  $10^7 \text{ cm}^2 \text{ s}^{-1}$ .

The model domain covers the Pacific Ocean from  $30^\circ\text{S}$  to  $50^\circ\text{N}$  with realistic coastal geometry. The model ocean has a constant depth of 4149 m. The longitudinal resolution is  $1^\circ$  uniformly. The latitudinal resolution is  $1/3^\circ$  within  $10^\circ\text{S}$  and  $10^\circ\text{N}$ , and gradually increases poleward to  $2.5^\circ$  at  $30^\circ\text{S}$  and  $50^\circ\text{N}$ . There are 27 levels throughout the water column with 10 uniform layers in the upper 100 m and 8 additional levels between 100 and 317 m. A rigid lid approximation is made at the surface where the vertical velocity is zero. The time step is 1 hour and 3-day averages of the temperature, salinity, and the three velocity components are stored.

The wind stress acts as a body force to the first layer of the model. Initial conditions of the OGCM temperature and salinity distributions were defined by the Levitus [1982] climatological-mean January distributions; the initial velocity was zero. During the first three years the model ocean is forced with the Hellerman and Rosenstein [1983] climatological-mean monthly surface

1 wind stress and the *Oort et al.* [1987] climatological-mean monthly surface air temperature.  
 2 Monthly mean oceanographic conditions during the third year are defined to be the climatological-  
 3 mean annual cycle and called climatology. The equatorial Pacific vertical motion annual mean  
 4 climatology, which indicated maximum upwelling at 50 - 100 m depths and maximum upwelling  
 5 near 180° - 150°W with diminished upwelling eastward and westward of this region [*Philander et*  
 6 *al.*, 1987], was substantiated by *Halpern et al.* [1989] and *Halpern and Freitag* [1987].

7 Simulated oceanographic conditions at the end of the third year become the initial conditions  
 8 for two simulations of the 1987 El Niño and 1988 La Niña conditions. Why employ simulations  
 9 of the 1987 and 1988 oceanographic conditions when the CZCS sensor had already ceased to  
 10 transmit data? Equatorial oceanic flow and thermal fields are very responsive to variations in the  
 11 surface wind field, which was not as well determined in 1982 and 1983 as it was in 1987 and  
 12 1988. Winds during an El Niño episode have some common features, such as the reduction of the  
 13 westward component. The longitudinal extent of the reduction does vary from one El Niño to  
 14 another. We believe that the simulated oceanographic conditions for 1987 are a reasonable  
 15 representation of the 1982 - 1983 El Niño episode because the 1987 minimum value of the  
 16 Southern Oscillation, which is an important index of the strength of El Niño, was only eclipsed in  
 17 1982-83 during the past twenty years [*Climate Diagnostic Bulletin*, 1992].

18 Two readily available wind products for 1987 and 1988 were used. The OGCM was forced  
 19 with the surface wind stress calculated from monthly-mean pseudo-stress data [*Goldenberg and*  
 20 *O'Brien*, 1981] acquired from the Florida State University (FSU) and with the surface air  
 21 temperature acquired from the European Center for Medium-Range Weather Forecasting  
 22 (ECMWF); this simulation was named FSU. In the second simulation of 1987 El Niño and 1988  
 23 La Niña conditions, the OGCM was forced with ECMWF surface wind stress; this simulation was  
 24 named ECMWF. A drag coefficient of  $1.4 \times 10^{-3}$  was used with both the FSU pseudo-stress  
 25 components and ECMWF 10-m height wind speed components [*Chao et al.*, 1992]. The FSU  
 26 wind stress was about 30 - 40% greater than that of ECMWF [*Chao et al.*, 1992].

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### 3. RESULTS

#### 3.1. CZCS Data

Only five time intervals of longitudinal distributions of 1°S - 1°N CZCS-derived phytoplankton pigment concentration were assembled along the equator from 145°E to 95°W (Figure 1): November 1979 - January 1980, May 1981, November 1981 - January 1982, May 1982, November 1982. The areal coverage of the GSFC-archived 2-minute satellite-duration CZCS images measured approximately 2000 km in longitude and 1000 km in latitude at the equator. The number of CZCS images used herein was 525 (Table 1), which was about 0.4% of the total number of images recorded worldwide throughout the 1979 - 1986 CZCS mission [McClain *et al.*, 1993]. The minimum number of 2-minute images required for a monthly composite phytoplankton pigment distribution along the equator from 145°E to 95°W was seven, which was approximately one-eighth of the monthly number of CZCS images (Table 1).

Data recorded during the two November - January intervals were combined, similarly for the two May periods. Data gaps occurred in different locations along the equator in each November - January and May interval, coincident data were averaged, and the 0.1 mg m<sup>-3</sup> spikes near 175°E and 175°W (Figure 1A) occurred in only one time period. Because of the absence of frequently sampled CZCS data during a month, individual spikes are probably not representative of the monthly mean condition, and attention is focused upon CZCS trends along the equator. The El Niño condition was represented by November 1982 when the El Niño intensity was very strong along the equator [Halpern, 1987]. Except for November 1982, each monthly or seasonal composite is considered to represent climatological-mean conditions.

The longitudinal-averaged phytoplankton pigment concentrations were 0.13, 0.10, and 0.15 mg m<sup>-3</sup> during November - January season, (Figure 1A), November 1982 during the El Niño (Figure 1B), and during May (Figure 1C), respectively. That no values greater than 0.35 mg m<sup>-3</sup> occurred from 145°E - 95°W in the monthly and seasonal composites (Figure 1) indicated that the



1 open-ocean equatorial Pacific had lower phytoplankton biomass than coastal waters [*Feldman*,  
 2 1986], including the Galapagos Islands [*Feldman et al.*, 1984]. In comparison, a phytoplankton  
 3 pigment concentration of  $0.7 \text{ mg m}^{-3}$  is the minimum value necessary to satisfy the carbon  
 4 requirement of anchovy larvae in the Peruvian coastal upwelling system [*Walsh et al.*, 1980].

5 The zonally averaged, monthly (or seasonally) mean euphotic-layer depth, which was  
 6 estimated from the CZCS-derived phytoplankton pigment concentration (Figure 1D) and the *Morel*  
 7 [1988] model and which corresponds to the 1% level of photosynthetically available radiation  
 8 (PAR), was about 90 m throughout the length of the Pacific equator. In §4 the vertical advection  
 9 of nitrate at 90 m, *i. e.*, into the euphotic layer, will be described. This depth was equal to  
 10 approximately three times the Secchi disk depth [*Lewis et al.*, 1988]. Using PAR\* measurements,  
 11 *Wilkerson and Dugdale* [1992] determined a 62-m 1% light penetration depth at  $0^\circ$ ,  $150^\circ\text{W}$  during  
 12 3 - 6 March 1988. *Fiedler et al.* [1992] reported that the August - November  $130^\circ\text{W}$  -  $95^\circ\text{W}$   
 13 equatorial euphotic-layer depths were 60 - 70 m in 1987 and 50 - 60 m in 1988. During a 23  
 14 March - 8 April 1992 time series of 1% PAR measurements at  $0^\circ$ ,  $140^\circ\text{W}$  the average euphotic-  
 15 layer depth was 85 m (*C. Davis*, personal communication 1992).

16 Because minimum and maximum westward wind speeds along the equator occur in April and  
 17 October, respectively, the annual cycle of the CZCS-derived phytoplankton pigment concentration  
 18 was portrayed by the May (Figure 1C) and November - January (Figure 1A) intervals. The May  
 19 and November - January longitudinal distributions were similar. Differences between the average  
 20 CZCS-derived phytoplankton pigment concentrations from  $145^\circ\text{E}$  -  $95^\circ\text{W}$ ,  $145^\circ\text{E}$  -  $170^\circ\text{E}$ ,  $175^\circ\text{E}$ -  
 21  $95^\circ\text{W}$ , and  $175^\circ\text{W}$  -  $150^\circ\text{W}$  were 0.02, 0.02, 0.01, and  $0.03 \text{ mg m}^{-3}$ , respectively. In each region  
 22 the approximate range of the annual cycle was less than the detectable resolution of CZCS data.

23 The annual mean longitudinal distribution of the CZCS-derived phytoplankton pigment  
 24 concentration (Figure 1D) was equal to the average of the November - January (Figure 1A) and  
 25 May (Figure 1C) values. The annual mean phytoplankton concentrations from  $145^\circ$  -  $165^\circ\text{E}$  and  
 26  $175^\circ\text{W}$  -  $95^\circ\text{W}$  were essentially uniform with mean  $\pm$  standard deviation values of  $0.07 \pm 0.01$  and  
 27  $0.15 \pm 0.01 \text{ mg m}^{-3}$ , respectively. The doubling in phytoplankton concentration from  $165^\circ\text{E}$  to

1 175°W was approximately linear which, also, was observed by *Barber and Chavez* [1991] from *in*  
 2 *situ* measurements. *Dandonneau* [1992] reported that the mean 1978 - 1989 surface chlorophyll  
 3 concentrations at the equator within 10° longitude of 165°W and 100°W were 0.12 and 0.15 mg  
 4 (chlorophyll equivalent) m<sup>-3</sup>, respectively. The two values were remarkably similar to the CZCS-  
 5 derived phytoplankton pigment concentrations (Figure 1D) and the 0.03 mg m<sup>-3</sup> difference  
 6 between the eastern and central Pacific was not significant because of the 0.037 mg m<sup>-3</sup> minimum  
 7 detection level of the CZCS sensor.

8 A single example of an interannual variation was defined by the CZCS difference between  
 9 November 1982 and the average November - January interval. Ocean biomass in the central  
 10 Pacific from 175°W to 155°W was reduced from 0.13 mg m<sup>-3</sup> (Figure 1A) to 0.05 mg m<sup>-3</sup> (Figure  
 11 1B) during El Niño; the decrease, which was more than 250%, was easily detectable by the CZCS  
 12 instrument. This dramatic variation in chlorophyll-*a* strongly influenced the marine bird population  
 13 at Christmas Island (2°N, 157°W) [*Schreiber and Schreiber*, 1989]. During El Niño, the  
 14 phytoplankton biomass from 145°E to 170°E in the western Pacific grew by nearly 50% from 0.07  
 15 mg m<sup>-3</sup> (Figure 1A) to 0.10 mg m<sup>-3</sup> (Figure 1B), which was likely aided by the uplifted  
 16 thermocline and smaller mixed layer depth known to occur in the western Pacific during El Niño.  
 17 Four months after November 1982 the marine resources near the Galapagos Islands were also  
 18 severely impacted by El Niño related changes in the phytoplankton concentration [*Feldman et al.*,  
 19 1984]. Throughout the Pacific equator, the magnitude of the 1982 El Niño interannual variation  
 20 was greater than the annual cycle.

21 Within 5° of the equator (equivalent to the distance of 3 or 4 times the Rossby radius of  
 22 deformation) the CZCS-derived phytoplankton pigment concentration at the Pacific equator was  
 23 larger than that towards the north and south (Table 2). The north-south distribution was  
 24 approximately symmetrical about the equator. Ratios of the equatorial phytoplankton pigment  
 25 concentration and the average of the 5°S and 5°N values, which were almost the same, were  
 26 computed at 10°-longitudinal intervals between 160°E and 110°W. All ratios were greater than  
 27 unity. The zonal average ratio was 1.5. Ratios greater than 1.5 occurred from 140°W to 110°W,

1 which were similar to that computed from the *Peña et al.* [1990] and *Chavez et al.* [1990] *in situ*  
 2 measurements; ratios less than 1.5 occurred west of 140°W.

3 That the phytoplankton biomass was greater at the equator than at 5° latitude is illustrative of  
 4 the biological "greening" of a narrow region along the Pacific equator. Considerable evidence  
 5 supports the *Cromwell* [1953] suggestion that equatorially-trapped physical processes of upwelling  
 6 and mixing enable more nutrients to reach the euphotic layer at the equator than at 5° latitude [e. g.,  
 7 *Halpern and Barber*, 1986; *Betzer*, 1992]. The meridional maximum in biological activity has been  
 8 noted by oceanographers for decades and has recently been brought to the public's attention by  
 9 publication of G. Feldman's color-coded displays of the global CZCS phytoplankton pigment  
 10 concentration [e. g., *U.S. News & World Report*, 1989].

### 12 3.2. Vertical Velocity Simulations

14 One linkage between vertical velocity and phytoplankton concentration is by vertical  
 15 advection of nitrate into the 90-m thick euphotic layer, where adequate light permits phytoplankton  
 16 growth to occur. Thus, the 90-m vertical velocity is emphasized.

17 Consider the average of the 1987 El Niño and 1988 La Niña simulations to be representative  
 18 of the climatological-mean condition because extreme opposite environmental situations occurred  
 19 in 1987 and 1988. Along the Pacific equator from 145°E to 95°W, the mean 1987 - 1988  
 20 longitudinal-averaged FSU-simulated vertical motion at 90 m was  $1.0 \times 10^{-5} \text{ m s}^{-1}$  (or 88%) greater  
 21 than that produced with the ECMWF wind (Figure 2). The 2-year averaged FSU and ECMWF  
 22 simulations are compared with climatology. Throughout the 145°E - 95°W region, the FSU-  
 23 simulated upwelling intensities along the equator were nearly identical to that of climatology  
 24 (Figure 2). In contrast, only in the western and eastern regions were ECMWF-simulations equal  
 25 to that of climatology (Figure 2). From 180° - 115°W the ECMWF-simulated upwelling was  
 26 considerably smaller than the climatological-mean simulation and, unlike the FSU simulation, did  
 27 not portray a definitive maximum in the central Pacific, which was shown to exist by *Halpern et al.*

[1989]. Thus, ECMWF-simulated vertical velocity was not considered to be as accurate as the FSU-simulated vertical motion, and the FSU-simulated vertical motion was used for all succeeding calculations.

#### 4. NITRATE ADVECTION

In the equatorial Pacific, nitrate,  $\text{NO}_3$ , is the dominant inorganic species used for new production by phytoplankton [Dugdale *et al.*, 1992]. The idea of using temperature to estimate nitrate in upwelling regions is not new [e. g., Dugdale *et al.*, 1989]. The average relationship between temperature and nitrate concentration for temperatures from 20 °C to 30 °C in the equatorial Pacific region from 180° to 93°W and 1.1°S to 1.1°N was nearly linear:

$$\text{NO}_3 [\mu\text{M}] = 34.75 - 1.125 \text{ Temperature } [^\circ\text{C}] \quad (3)$$

Equation (3) was computed from a scatter diagram prepared by S.T. Lindlay [personal communication, 1992] from the data set described by Barber and Chavez [1991]. Temperatures above 30.8 °C, which correspond to zero nitrate, were arbitrarily set to 30.8 °C, and Equation (3) was also used for temperatures as low as 13°C. Equation (3) numerical values were each approximately 60% smaller than that determined by Fielder *et al.* [1991] for the eastern tropical Pacific from about 5°S to 20°N and from 130°W to the American continent.

Monthly mean nitrate concentrations at 85 and 95 m were computed from the FSU-simulated monthly mean temperatures at 85 and 95 m, according to Equation (3). The nitrate determined from the simulated temperature increased with depth. The FSU-simulated 85- and 95-m nitrate during March 1988 at 0°, 150°W were 5.3 and 6.2  $\mu\text{M}$ , respectively. The FSU-simulated 90-m nitrate was about one-half smaller than the Wilkerson and Dugdale [1992]  $\text{NO}_3$  *in situ* measurements at 0°, 150°W during 2 - 7 March 1988. It is not unreasonable for the 5-day averaged value of *in situ* data to be two times different than a monthly mean value because of the very large amplitudes of submonthly fluctuations which occur along the Pacific equator [Halpern *et al.*, 1988]. The absolute values of the differences between the 60-m climatological-mean annual

1 simulated nitrates along the equator between 145°E - 180°, 180° - 140°W, and 140°W - 100°W and  
 2 the *Barber and Chavez* [1991] 60-m *in situ* nitrate measurements averaged between 5°S to 5°N  
 3 during 16 expeditions from 1983 to 1990 were 0.6, 2.1, and 0.8  $\mu\text{M}$ , respectively. The agreement  
 4 was excellent, except in the central Pacific where *Barber and Chavez* [1991] reported a secondary  
 5 maximum from 160°W to 140°W, which was not simulated by the model. The August - November  
 6 1987 and 1988 FSU-simulated vertical averaged nitrate concentrations in the euphotic zone  
 7 between 130°W - 95°W were 3% and 11%, respectively, greater than that measured by *Fiedler et*  
 8 *al.* [1992]. The simulated climatological-mean annual vertical averaged euphotic-layer nitrate  
 9 decreased approximately linearly from 95°W to 170°W and was uniform to within  $\pm 0.5 \mu\text{M}$   
 10 westward of 170°W, which was in excellent agreement with the pattern inferred from the depth-  
 11 longitude 5°S - 5°N section reported by *Barber and Chavez* [1991]. Thus, numerical simulations  
 12 of nitrate appeared sensible.

13 The linear vertical gradient of nitrate at 90 m was calculated from the 85- and 95-m nitrate  
 14 values:

$$15 \quad \partial\text{NO}_3/\partial z = ((\text{NO}_3)_{85\text{m}} - (\text{NO}_3)_{95\text{m}})/10 \quad [\mu\text{M m}^{-1}] \quad (4)$$

16 The longitudinal distributions of the average 1987 - 1988 FSU and annual-mean climatology  
 17 simulations of the vertical nitrate gradients were almost the same (Figure 3), except in the far  
 18 western Pacific. Maxima ( $\approx -0.035 \mu\text{M m}^{-1}$ ) and minima ( $\approx -0.085 \mu\text{M m}^{-1}$ ) occurred near 180°  
 19 and 125°W, respectively. Zonal average values of the FSU and climatology simulations differed  
 20 by 7%.

21 Advection of nitrate into the euphotic zone was equal to the product of the vertical velocity,  
 22  $w$ , and  $\partial\text{NO}_3/\partial z$ . A negative value means that the water parcel moves upward into a region of  
 23 lower nitrate or moves downward into a region of greater nitrate; since the 90-m  $w$  was positive  
 24 virtually everywhere from 145°E to 95°W (Figure 2), the former situation applies. At 150°W, the  
 25 March 1988 90-m FSU-simulated vertical flux was about  $-40 \times 10^{-7} \mu\text{M s}^{-1}$ , which was nearly 40  
 26 times greater than the nitrate uptake rate by phytoplankton determined by *Dugdale et al.* [1992]  
 27 from  $^{15}\text{NO}_3$  measurements during 2 - 7 March 1988. This severe imbalance between the nitrate

flux into the euphotic zone and the assimilation rate of nitrate by phytoplankton is an important feature of the enigma why the equatorial ocean is not greener than it is. During March 1988, the 90-m FSU-simulated vertical flux (not shown) was maximum at 0°, 150°W and not representative of the zonal distribution. At 155 and 145°W the March 1988 90-m nitrate fluxes were each about  $-20 \times 10^{-7} \mu\text{M s}^{-1}$ , which was approximately equal to the zonal average of  $-17 \times 10^{-7} \mu\text{M s}^{-1}$ . The unused nitrate upwelled at the equator is probably advected poleward by the Ekman drift current, which could account for the nearly uniform near-surface nitrate distribution from 5°S to 3°N [Peña *et al.*, 1990].

## 5. RELATIONSHIP OF UPWELLING AND PHYTOPLANKTON BIOMASS

### 5.1. Annual Mean

Very little resemblance was found between the bell-shaped longitudinal pattern of upwelling (Figure 2) and the two-horizontal line pattern of the phytoplankton pigment concentration (Figure 1D), except for the linear increase in the 165°E - 175°W region. The similar 165°E - 175°W shapes of the upwelling and phytoplankton pigment distributions are presumed to be coincidental. It is tempting to speculate that the 165°E - 175°W linear gradient of phytoplankton biomass was associated with the annual east-west movement of the western edge of the cold tongue, which would be associated with a higher phytoplankton concentration.

Zonal distributions of the average 1987-88 FSU and climatology simulated vertical flux of nitrate were remarkably similar, with each pattern having a minimum near 130°W (Figure 4). The 145°E - 95°W average values were equivalent to 10%.

Little if any direct correspondence was observed between the longitudinal distributions of the amount of CZCS-derived pigment concentration (Figure 1D) and nitrate flux into the euphotic layer (Figure 4). From 175°W to 130°W the upward nitrate flux increased four times from 7 to 27  $\mu\text{M s}^{-1}$  while phytoplankton concentrations remained essentially uniform. Likewise, from 130°W to

110°W the upward nitrate flux decreased by a factor four, while the phytoplankton content remained constant. The computed longitudinal distribution of the upward flux of nitrate was related to the nitrocline depth, which, like the thermocline, decreases from west to east. If the nitrocline depth was shallow (e. g., at 95°W) or deep (e. g., at 175°W) relative to 90 m, then the computed nitrate gradient and upward flux of nitrate at 90 m depth would both be small compared to the location where the nitrocline was near 90 m (e. g., near 130°W - 120°W).

## 5.2. Annual Cycle

The 1987 - 1988 FSU and climatology simulations of the May (Figure 5A) nitrate flux distributions were similar; also, for November - January (Figure 5B). The May nitrate flux fluctuations occurring east of 160°W with an approximate 1000-km zonal wave length were probably manifestations of westward propagating 20- to 30-day period mixed Rossby-gravity waves [Halpern *et al.*, 1988]. The waves were not as prominent in the average November - January pattern because the averaging process yields a reduced apparent amplitude.

The annual cycle range of the phytoplankton pigment concentrations (Figure 1A and 1C) was probably too small to be detected. The 145°E - 95°W zonal average phytoplankton pigment concentration in May was 14% greater than that during November - January.

The annual cycle range of the 90-m nitrate flux (Figure 5) was almost nonexistent because the May and November - January nitrate flux simulations were very similar. The 145°E - 95°W zonal average upward flux of nitrate in May was 6% greater than that during November - January. The longitude of the minimum nitrate flux shifted eastward from about 130 - 125°W in May to near 125°W in November - January.

The relationship between May and November - January longitudinal profiles of phytoplankton content and nitrate flux was not *a priori* expected. The 145°E - 95°W zonal average climatology simulated upwelling intensity was  $0.3 \times 10^{-5} \text{ m s}^{-1}$  (or 15%) weaker in May than during November - January, which is consistent with reduced easterly wind speed in May compared to

November - January. If upwelling of nitrate into the euphotic zone was a prominent growth mechanism of phytoplankton, then the phytoplankton biomass and upward nitrate flux were expected to be smaller in May than during November - January. There was no difference in calculated phytoplankton biomass for these two periods.

### 5.3. *El Niño Variation*

The 175°W - 155°W phytoplankton pigment concentration was 250% smaller in an El Niño November (i. e., November 1982 (Figure 1B)) than during a typical November - January interval (Figure 1A). El Niño variations of phytoplankton biomass were consistent with simulations of the flux of nitrate into the euphotic layer. In November 1987, when an El Niño occurred, the average 175°W - 155°W FSU-simulated 90-m nitrate flux into the euphotic layer was 60% less than that prescribed by the climatology simulation. Thus, it is tempting to speculate that the reduced phytoplankton biomass during El Niño was caused by a decrease in upward flux of nitrate into the euphotic layer. Eastward advection of high temperature, low plankton biomass water, which normally resided west of 165°E (Figure 1A), could also contribute to the reduced phytoplankton concentration from 175°W - 155°W in November 1982 (Figure 1B).

During the November 1988 La Niña conditions, which are analogous to conditions that are opposite of El Niño, the average 145°E - 95°W upward nitrate flux was 60% greater than the November climatology and 70% greater than that in November 1987. The local upwelling influence seemed particularly important because the 175°W - 155°W average FSU-simulated upwelling speed in November 1988 was  $7.8 \times 10^{-5} \text{ m s}^{-1}$  compared to  $1.1 \times 10^{-5} \text{ m s}^{-1}$  in November 1987. Unfortunately, there was not sufficient sampling of CZCS data in 1983 - 1984 to describe La Niña conditions.

## 6. SUMMARY AND DISCUSSION



1       The CZCS-derived measurements of phytoplankton pigment concentration along the Pacific  
 2 equator were analyzed *vis-a-vis* longitudinal pattern, annual cycle, and El Niño variations. The  
 3 small amplitude of the annual cycle was almost undetectable. The 1982 El Niño variation of  
 4 phytoplankton biomass in the central Pacific was substantial.

5       The simulated monthly mean euphotic-layer nitrate along the equator was in very good  
 6 agreement with observations reported by *Barber and Chavez* [1991] and *Wilkerson and Dugdale*  
 7 [1992]. The simulated upward flux of nitrate into the euphotic layer provided more than enough  
 8 nitrate for assimilation by phytoplankton, and was further evidence for the importance of *Barber's*  
 9 [1992] question, "Why isn't the equatorial Pacific greener?" Two reasons why the phytoplankton  
 10 abundance is less than the amount expected from the high nutrient concentration are grazing and  
 11 iron limitation [*Frost and Franzen*, 1992].

12       Upwelling at the equator provided a continuous upward flux of nitrate. That all along the  
 13 equator the CZCS-derived phytoplankton pigment concentration was maximum at the equator  
 14 compared to 5° latitude attests to the importance of upwelling. The influence of the upwelling of  
 15 nitrate into the euphotic layer was examined. The annual mean longitudinal profiles of the CZCS-  
 16 derived phytoplankton pigments and the 90-m nitrate flux bore no resemblance to each other. The  
 17 very small amplitudes of the annual cycles of the phytoplankton and nitrate flux had the same  
 18 phase: a larger biomass was associated with a larger upward flux, and *vice versa*. However, the  
 19 increased nitrate flux was not expected in May because the intensity of wind-driven upwelling was  
 20 smaller then than in November - January.

21       El Niño variations of phytoplankton pigment concentration were in phase with the 90-m  
 22 nitrate flux. In the central Pacific (where there were sufficient data) the phytoplankton biomass and  
 23 nitrate flux were both reduced during El Niño compared to non-El Niño conditions. In November  
 24 1987 during the waning stage of an El Niño the smaller upward flux of nitrate, which was  $-4 \times 10^{-7}$   
 25  $\mu\text{M s}^{-1}$  from 175°W - 155°W, could be surmised to have also occurred in November 1982 during  
 26 the onset phase of an El Niño and, therefore, contributed directly to the reduced phytoplankton  
 27 abundances, which were  $0.05 \text{ mg m}^{-3}$  (Figure 1B).

1        In conclusion, the objective to describe annual and interannual variations of the CZCS-  
2        derived phytoplankton pigment concentration was severely limited by the scarcity of time series  
3        measurements. We could not explicitly define to what extent upwelling controlled phytoplankton  
4        abundance along the Pacific equator. With the scheduled launch of the satellite-borne Sea-viewing  
5        Wide-field-of-view Sensor (SeaWiFS) in August 1993, ocean color measurements will become  
6        continuous in space and time in the equatorial Pacific.

7  
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1 Table 1. Monthly number of CZCS images used to compute the phytoplankton pigment  
2 concentrations displayed in Figure 1.

3	Month	Number of Images
4	Nov 1979	47
5	Dec 1979	60
6	Jan 1980	42
7	May 1981	30
8	Nov 1981	81
9	Dec 1981	85
10	Jan 1982	118
11	May 1982	38
12	Nov 1982	24



1 Table 2. The 1979 - 1986 averaged CZCS-derived phytoplankton pigment concentrations at 5°S,  
 2 0°, and 5°N at 10° intervals between 160°E and 110°W.

3				
4	Longitude	5°S	0°	5°N
5		mg m <sup>-3</sup>	mg m <sup>-3</sup>	mg m <sup>-3</sup>
6	110°W	0.090	0.160	0.085
7	120°W	0.085	0.129	0.085
8	130°W	0.071	0.145	0.010
9	140°W	0.076	0.138	0.080
10	150°W	0.080	0.120	0.086
11	160°W	0.070	0.112	0.090
12	170°W	0.088	0.104	0.071
13	180°	0.084	0.110	0.068
14	170°E	0.056	0.090	0.066
15	160°E	0.052	0.062	0.051

# LIST OF FIGURES

1

2

3 Figure 1. Zonal distributions of CZCS-derived phytoplankton pigment concentrations along the  
 4 Pacific equator for periods listed in each panel. The average May 1981 and May 1982 composite  
 5 contained single values of 1.3 and 0.47 mg m<sup>-3</sup>, which were deleted so that 0.4 mg m<sup>-3</sup> became the  
 6 maximum value for each panel.

7

8 Figure 2. The average 1987 - 1988 and climatological-mean annual zonal distributions of 90-m  
 9 vertical velocity simulated with FSU and ECMWF winds during 1987 and 1988 and with  
 10 Hellerman and Rosenstein (1983) climatological winds.

11

12 Figure 3. The average 1987 - 1988 and climatological-mean annual zonal distributions of the  
 13 vertical gradient of nitrate at 90 m.

14

15 Figure 4. The average 1987 - 1988 and climatological-mean annual zonal distributions of the  
 16 vertical flux of nitrate at 90 m.

17

18 Figure 5. The average 1987 - 1988 and climatological-mean zonal distributions of the vertical flux  
 19 of nitrate at 90 m: (A) May and (B) November - January.

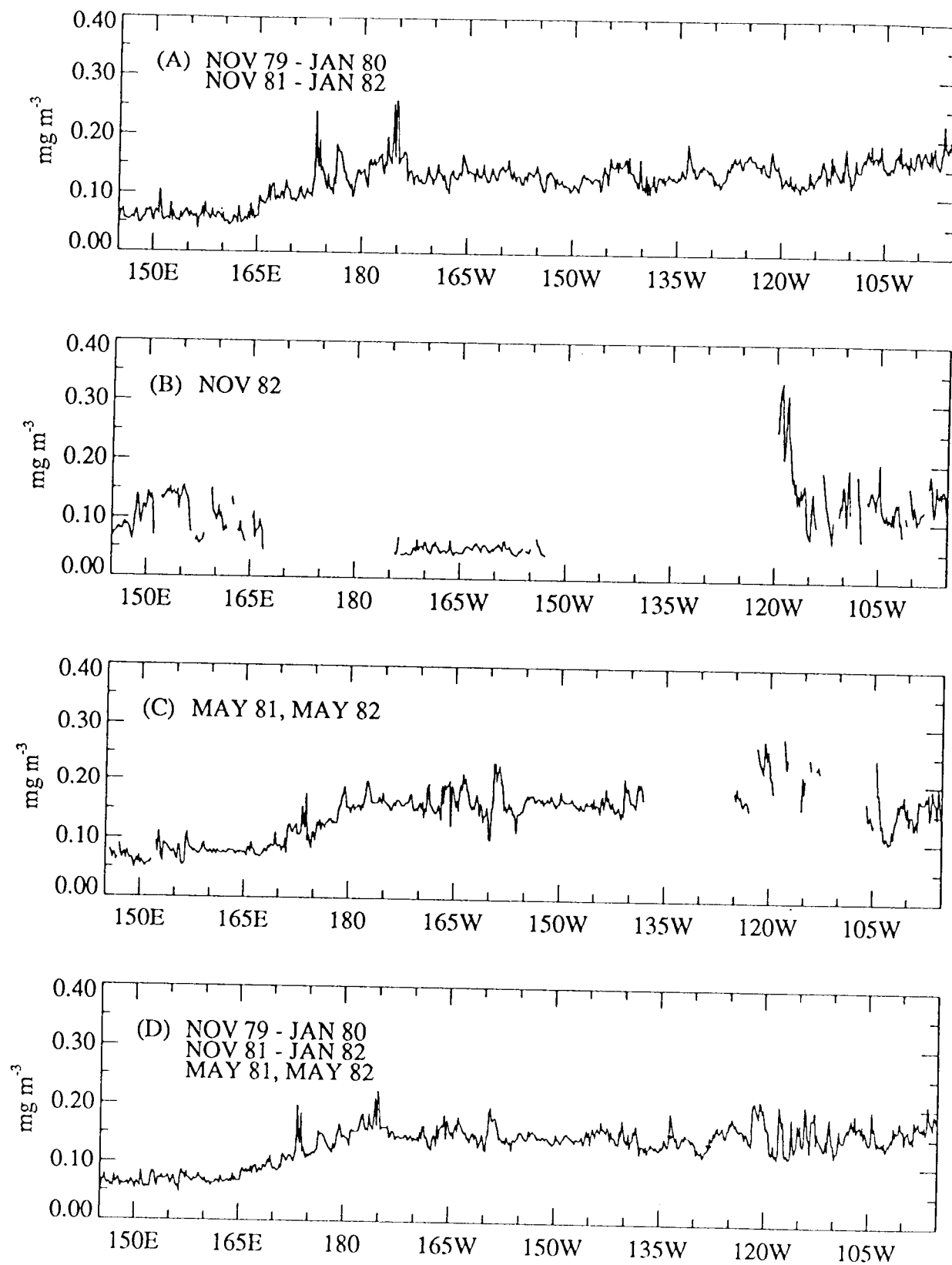


Figure 1

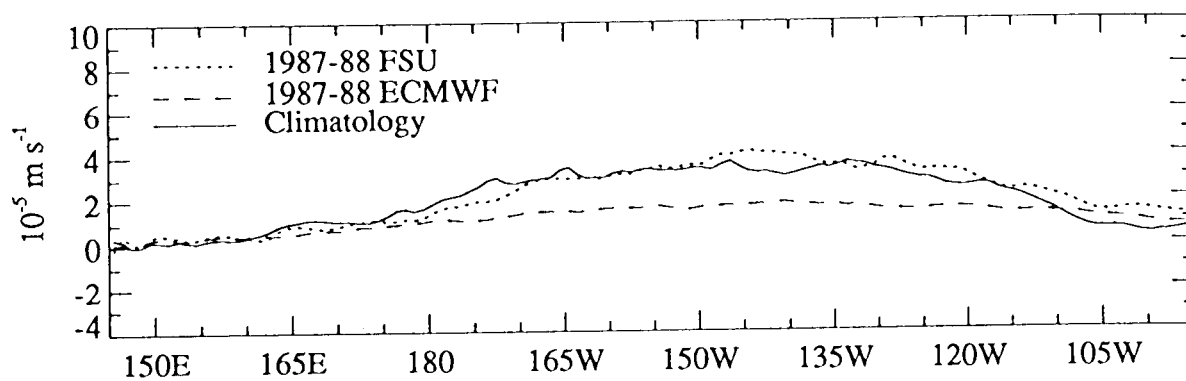


Figure 2

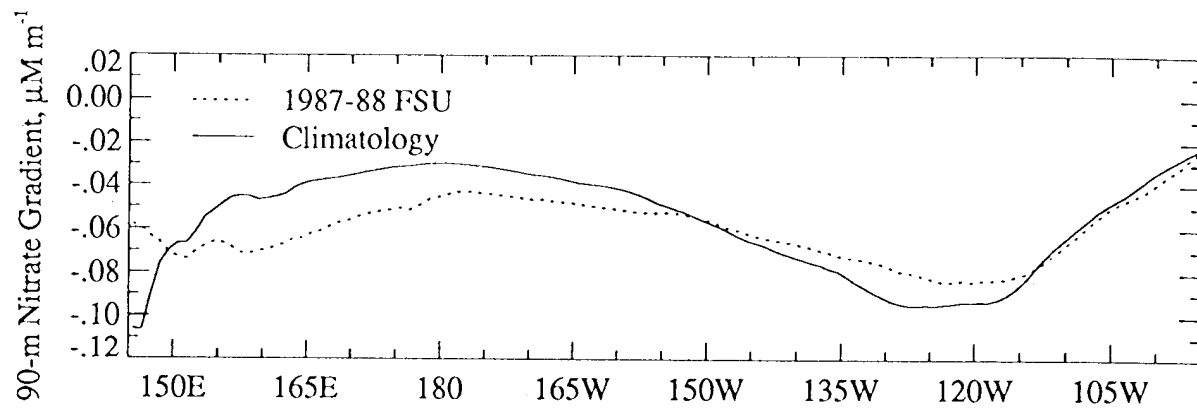


Figure 3

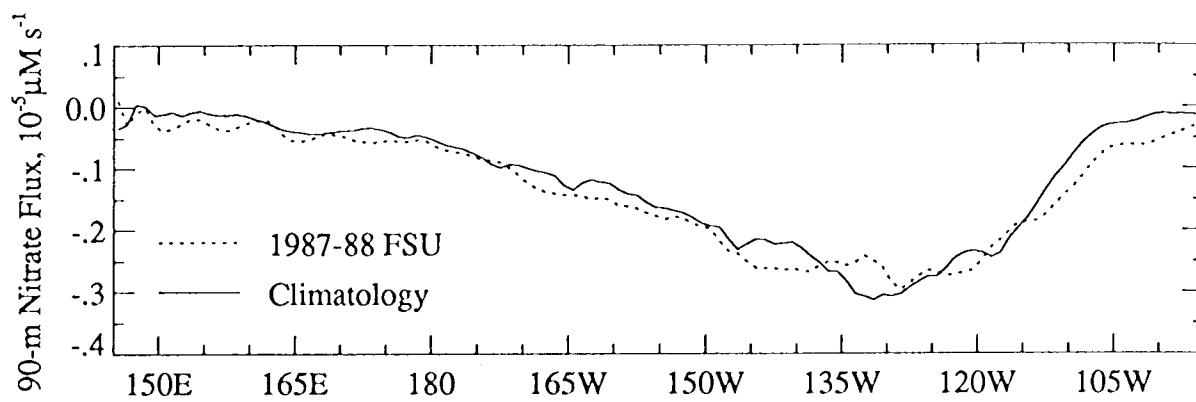


Figure 4